

# LONG-TERM SPATIALLY DISTRIBUTED SOLUTIONAL EROSION: HOW DO WE PUT SOLUTES ON THE MAP?

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## ABSTRACT

Much of the short-term solute process research in the last 20 years has focused on hydrochemical behaviour. This can help with the identification of solute source area type, but not the actual location of the solutional loss. The location and the spatial distribution of erosion are crucial to the identification of long-term landscape evolution. The way ahead lies in the fact that catchment budgeting and geochemical studies have been able to provide solute uptake rates. If these can be related to soil type, through a knowledge of soil mineralogy, reactive surface area and water flow rate, it may be possible to characterize the location and spatial distribution of solutional erosion through terrain analysis and soil maps in a geographical information system. © 1997 by John Wiley & Sons, Ltd.

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## INTRODUCTION

During the 1970s, British geomorphology moved away from attempts to reconstruct landscape evolution largely on the basis of landform evidence, coupled with Davis's theoretical model of 'normal' landscape evolution, towards a focus on the relationship between process and landforms. This led to a large number of studies attempting to relate processes to landforms, and to the development of process–response models which attempted to model landforms on the basis of process (e.g. Carson and Kirkby, 1972; Burt, 1986; Crabtree, 1986; Trudgill, 1986a).

In the field of solute studies, early work on monitoring the behaviour of solutes in streams was concerned, at least in part, with what this could tell us about the nature and location of solutional processes on the slopes (e.g. Lewin *et al.*, 1974), but later work has seen a change in emphasis towards the monitoring and modelling of solute concentrations in the water *per se*, often in applied contexts such as studies of acidification or nutrient loading from agriculture (Trudgill, 1995).

Despite the fact that many of the studies were never intended to elucidate long-term landform evolution, it is pertinent to ask what use the years of solute monitoring and measurement have been to the understanding of landforms. Three threads of discussion have been identified in Figure 1.

## HYDROCHEMICAL BEHAVIOUR

Solute dynamics have been well studied since the time of Johnson *et al.* (1969), and summarized by Gregory and Walling (1973) and Trudgill, (1986b, 1995), but the behaviour of solutes in relation to discharge over time in effect tells us little about landform evolution. Approaches such as end-member mixing analysis (EMMA; Wheater and Beck, 1995) can enable the identification of source areas with distinctive chemistries arising from soil, hillslope and groundwater sources (the first often being higher in nitrate and sulphate and the last being higher in silica in many examples). However, this only tells us about source area type rather than source

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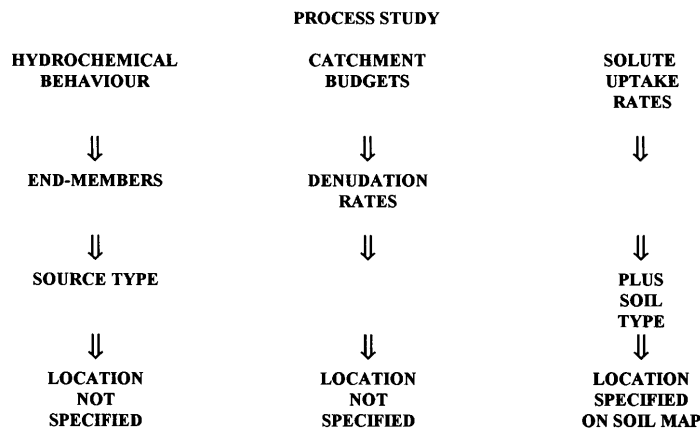


Figure 1. Relationship between short-term solute studies and long-term landform evolution.

location or spatial distribution. Distributed, physically based models attempt to do the same and could provide spatial solute rates, but since the model testing is aspatial (i.e. via the chemography) it is hard to rely on their conclusions. We can thus identify soil water but cannot distinguish between whether the primary source is from soils at the slope crest or the slope foot, or whether the solutes are derived uniformly or progressively down the slope.

### CATCHMENT BUDGETS

A significant development has been the ability to estimate catchment denudation rates through the year-long monitoring of solute concentrations and discharge and the use of rating curves and flow durations to derive catchment output values in  $\text{kg ha a}^{-1}$  or  $\text{mol m}^{-2} \text{a}^{-1}$ . Given appropriate budget adjustments for atmospheric inputs and biological factors (Velbel, 1995), the output from the catchment is equivalent to the weathering rate (Goudie, 1995).

The difficulties lie in the fact that we do not know how to distribute these losses either vertically, within the soil–bedrock profile, nor spatially across hillslopes. This is especially so since the losses are expressed per unit catchment surface area, whereas in fact they are derived from a three-dimensional particle surface area within the soil–bedrock profile.

### SOLUTE UPTAKE RATES

Unless the rates are derived from small topographic areas, their utility lies in providing comparative data for the rates of dissolution derived in the geochemical laboratory. Field rates are generally lower than laboratory rates, especially since many laboratory experiments remove the weathering products. Furthermore, laboratory studies and field evidence (Trudgill *et al.*, 1996; Sverdrup and Warfvinge, 1988) have suggested that a key determinant of stream water chemistry is the pathway of the water down the hillslope, and some studies have concluded that it is the characteristics of the riparian hillslope element which determines the solute characteristics of the stream water (Billett and Cresser, 1992).

In terms of modelling solutional erosion, however, it is not stream water quality which is important, but the uptake of solutes at all stages on the hydrological pathway. In the model of Carson and Kirkby (1972) this was assumed to depend on:

1. the quantity of each oxide present in the soil
2. the flow of water through the soil
3. the saturated solubility of the oxide.

The third of these assumes that soil water reaches equilibrium with the soil chemistry sufficiently rapidly that it can always be treated as saturated. However, Trudgill *et al.* (1996) have shown that this assumption will not

always be met. Given the normal description of solute uptake over time by a first-order kinetic equation, it is possible to predict the time taken for a solution to reach saturation. Given information on water flow rates in the soil, this can be converted to an estimate of the distance the water travels before reaching saturation, using the following equation:

$$d_{\max} = 3[(C_{\max}/k_r)/0.002(A_w/v)] \quad (1)$$

where  $d_{\max}$  = distance to reach  $C_{\max}$  (m),  $C_{\max}$  = maximum concentration ( $\text{mol l}^{-1}$ ),  $k_r$  = dissolution rate ( $\text{mol m}^{-2} \text{s}^{-1}$ ),  $A_w$  = area of weathering ( $\text{m}^2 \text{m}^{-3}$ ),  $v$  = water flow rate ( $\text{m s}^{-1}$ ). This converts a description of the process rate into a description of its spatial distribution by substituting distance downslope for time elapsed. The factors which control this substitution are all effectively determined by soil type:  $C_{\max}$  and  $k_r$  are mineral-specific,  $v$  is soil texture-specific as is  $A_w$  (though it can also be adjusted for the proportion of mineral present in a soil, e.g.  $A_w \times 0.1$  for 10 per cent composition (Pacés, 1983)). The input data required are therefore a knowledge of soil mineralogy, percentage composition of soil minerals, and soil texture, in order to derive the area of weathering and soil water flow rates.

The value of  $A_w$  can be for the depth of soil if the bedrock is impermeable, or a compound figure for the reactive surface in the soil depth plus the likely reactive surface of the bedrock. For deeply permeable bedrock, such as chalk, only the soil-surface contact area liable to contribute to surface lowering need to be considered. Since the interest is in the long-term evolution of landforms, variations in flow paths and velocities under different antecedent conditions need not be considered, but simply mean annual velocities. Solute inputs from rainfall can be accommodated by modifying Equation 1 to include the term  $C_{\max} - I$ .

Trudgill *et al.* (1996) have shown the following points, given values of  $A_w$  and  $v$  appropriate to soil texture.

(i) For rapidly dissolving solutes ( $k_r$  greater than  $1 \times 10^{-6} \text{ mol m}^{-2} \text{s}^{-1}$ ), path length is not important since whatever the hydrological or soil texture conditions (however rapid the flow or coarse textured the soil with low  $A_w$ ), they can reach  $C_{\max}$  at  $d_{\max}$  values of the order of  $1 \times 10^{-1}$  to  $1 \times 10^1 \text{ m}$ .

(ii) For most solutes, however slowly they dissolve, path length is not important in fine-textured soil as *high*  $A_w$ , greater than  $1 \times 10^5 \text{ m}^2 \text{m}^{-3}$ , and  $v$  less than  $1 \times 10^{-5} \text{ m s}^{-1}$  again means that they can reach  $C_{\max}$  at  $d_{\max}$  values of the order of  $1 \times 10^{-1}$  to  $1 \times 10^1 \text{ m}$ .

However, there are situations where  $d_{\max}$  is liable to be of the same order of hillslope length, for example with solutes dissolving at rates of less than  $1 \times 10^{-6} \text{ mol m}^{-2} \text{s}^{-1}$  in coarse-textured soils of  $A_w$  greater than  $1 \times 10^4 \text{ m}^2 \text{m}^{-3}$  and flows about  $1 \times 10^{-6} \text{ m s}^{-1}$ . Hence, given a hillslope element of length  $\lambda$  with uniform soil characteristics, there are three possible scenarios:

1.  $d_{\max} \ll \lambda$ , i.e. water reaches saturation very quickly
2.  $d_{\max} \gg \lambda$ , i.e. water never achieves saturation
3.  $d_{\max} \approx \lambda$ , i.e. slope length is critical.

In the first two cases, the original model of Carson and Kirkby is applicable; in the third case some modification will be necessary to allow for changing solute concentrations downslope.

The other simplifying assumption in the original model was that soil characteristics were uniform downslope. Although this is reasonable for a theoretical treatment, if we wish to relate actual process rates to the development of particular landscapes, the variation in soil characteristics must be included. This means that the length  $\lambda$  is not the length of the hillslope, but of each element of the slope with uniform soil characteristics.

Under conditions of scenarios 1 and 2 above, solute rates can be represented by areally characteristic values and plotted on a map as such using the suggestion of Drever (1988) from Lasaga (1984), converting  $k_r$  to  $k_e$  using the equation:

$$k_e = \{[(k_r m)/d]/s\} 10^{12} \quad (2)$$

where  $k_e$  = rate of retreat ( $\text{mm a}^{-1}$ ),  $k_r$  = dissolution rate ( $\text{mol m}^{-2} \text{s}^{-1}$ ),  $m$  = molecular weight,  $d$  = mineral particle density ( $\text{g cm}^{-3}$ ),  $s$  = number of seconds in a year (31 536 000) and the factor of  $10^{12}$  corrects the units for volume and  $\text{m}^2$  to  $\text{mm}$ .

Under conditions of scenario 3, where  $d_{\max}$  is less than the slope length of interest, surface lowering values will not be constant over space, and transport will not be limited by  $v$  but by slow values of  $k_r$ . Here, substitution of space for time through the intermediary of velocity is critical and the erosion rate will be a space–time–flow

dependent function of  $k_r$ ; the result will vary with flow rate and thus with texture. The result is thus best gained by considering flows in areal units of consistent soil texture (and therefore flow) and  $k_e$  values assigned to each soil unit.

The remaining problem is that  $k_e$  is for the mineral surface and not for the landform surface: the soil material does not become infinitely more porous during dissolution; mechanical settling and bulk density changes have to be allowed for if we are interested in surface lowering of the land. To this effect, it is evident that mechanical processes such as freeze–thaw, bioturbation and soil creep must be important mechanisms which actually intermediate between the dissolution of minerals and surface lowering. This can be allowed for by assuming that soil bulk densities and porosities remain constant over time, despite mass loss of solutes. Volume loss then becomes equivalent to mass loss divided by particle density and volume loss can be converted to surface lowering per unit area. Thus:

$$k_e = k_r/d \quad (3)$$

Selection of an appropriate areal unit suggests that  $\text{m}^2$  is more appropriate than  $\text{cm}^2$  or  $\text{km}^2$ , which are often used, and so conversion of units will be involved as particle densities are usually expressed in  $\text{g cm}^{-3}$  ( $=\text{Mg m}^{-3}$ ). Thus  $k_e = \text{m surface lowering per m}^2$ ,  $k_r = \text{solute loss in kg per m}^2$  ( $\text{kg ha}/10\,000$ ),  $d = \text{particle density in kg per m}^3$  ( $\text{g cm}^{-3} \times 1000$ ). Thus 1 mm surface lowering a year at  $d = 2.54 = 25.4 \text{ kg ha a}^{-1}$ , which can be applied to an areal unit provided the calculations above, which allow for  $A_w$  and  $v$ , indicate a constant value over space.

### IMPLEMENTATION

Although this is a theoretical discussion, it is worth speculating how the approach to modelling described here might be implemented. Two sorts of information will be required: (1) data on soil characteristics, particularly mineralogy and texture; (2) data on hillslope flow paths. The second is most easily derived using a digital elevation model (DEM) in a geographic information system (GIS). There are well established methods for deriving the direction of flow from each pixel in a gridded representation of the terrain (Moore *et al.*, 1993). The soil data can similarly be gridded, suggesting that it may be feasible to implement the entire model inside a GIS.

In such an implementation, the hillslope element length ( $\lambda$ ) is the portion of given flow path which has constant soil and flow characteristics. By combining the information on soil type, with some estimates of likely flow rates (simply related to soil texture, or incorporating topography using the topographic index of Beven and Kirkby (1979) it would be possible to partition a catchment into areas of uniform solute response (rather similar to the uniform hydrological units sometimes employed in runoff modelling). For each unit,  $d_{\text{max}}$  and  $\lambda$  could be calculated. Where these indicated scenarios 1 and 2, as described earlier, solute calculations could be done very simply for each pixel in the unit. In the case of scenario three, where flow path length is important, the calculations would have to take into account the distance downslope; again, this can be derived from a DEM using a GIS.

### EXTRAPOLATING TO THE LONG TERM

Most environmental modelling is concerned with predicting short-term process response in which topography can be regarded as constant (i.e. over the steady time-scale of Schumm and Lichty, 1965). In this case, however, the concern is with predicting changes in topography, which in turn will affect the distribution of soil types and the patterns of hillslope flow. This feedback must be built into the model, to avoid the simple-minded extrapolation of process rates.

It is well known that topography is a major control on soil formation; indeed, in a modelling exercise such as this it would be normal to hold most of the other factors, such as climate and parent materials, constant. Recent work by Moore *et al.* (1993) has shown that it is possible to predict soil characteristics with a reasonable degree of accuracy from topographic information alone. In their study of a 5.4 ha slope in Colorado, they showed that the topographic index could be used to predict the sand and silt contents of the soils with accuracies of 52 per cent and 64 per cent, respectively.

## CONCLUSIONS

Much of the work in measuring stream solutes has not been directly applicable to studies of landform evolution. However, a knowledge of the rates of solution under different soil types does allow us to identify those situations where solution is transport-limited, and those where path length will also be important. The combination of mapped data on soil type together with modelled flow rates and flow paths inside a GIS may therefore offer a promising means of using the results of process measurements for the modelling of landform evolution by solutional erosion.

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